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Yield Estimation and Mantle Q

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YIELD ESTIMATION AND MANTLE Q

By

Zoltan A. Der

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INTRODUCTION

In seismic methods of yield estimation the amplitudes of seismic waves originating from an explosion must be used to determine yields. The two basic types of seismic waves most frequently used are the teleseismic P waves and the long period Rayleigh waves of intermediate (20 sec) period. More recently a considerable amount of work has been done to utilize the amplitude measurements on Lg. For increasing the accuracy of yields some estimate of the attenuation of the wave types in question is necessary. Attenuation of 20 sec Rayleigh waves, according to the available evidence, does not vary much from region to region since most of the energy is contained in the high Q lithosphere, although discontinuities in the lithosphere may cause problems in some areas. The attenuation of Lg is mostly controlled by random heterogeneities in the crust (Aki 1969) and will not be discussed here. This presentation will be devoted to the attenuation effects on teleseismic, short period body waves.

Before discussing the results of the research concerning yield estimation from teleseismic P waves I want to touch on some special problems with estimating Q in the short period band that delayed progress in this field for a long time. A more detailed discussion on this topic by Cormier (1982) can be found in a recent issue of the BSSA.

The main methods for measuring Q in the short period band can be classified as follows:

a) Relative spectral ratio method, which cancels the source effects for axisymmetric sources at various stations.

b) Absolute modelling of the body wave spectra assuming some "plausible" source model.

c) Time domain modelling of the body wave waveforms using a suitable source time function.

The parameters of the path, the Q especially, must be adjusted to fit the observations in either the time or frequency domains with a frequency dependent or independent Q with some source functions. In the absence of noise all methods should give the same result. In practice, however, none of these conditions are met and unless some provisions are made to reduce the sensitivity of the method used for the estimation of Q to the numerous unknown parameters no meaningful results can be obtained. In this presentation I shall emphasize the spectral methods at the expense of time domain modelling of narrow-band waveforms and furnish facts that indicate that such methods are more robust, and less subject to bias due to the various still unresolved basic problems concerning seismic source and site related distortions of the spectra and the lack of the precise knowledge about the possible frequency dependence of Q.

In Slide 1 some theoretical shapes of P wave spectra are shown computed for a 200 kt explosion source using the VSB source model. The sensitivity of the shapes to t* is quite evident and small differences in t* are quite discernible in the t* range of 0-.3. At higher t* values the accuracy of t* measurement must decrease because of the limited frequency range available for the fit and the sensitivity of the details in the spectra to other perturbing factors.

The key to the estimation of the attenuation is the extreme sensitivity of the spectral shapes to Q, at the expense of other factors. Assuming losses in shear deformation in the mantle, which appears to agree with most experimental facts, and using some basic physical limits on plausible source functions it is possible, using spectral analyses, to put some limits on the possible range of Q and its frequency dependence.

To illustrate the insensitivity of short period P wave spectra to near-site focusing effects one can compare the variability of the P wave amplitudes, waveforms and shapes of spectra at seismic arrays. While seismic waveforms and amplitudes exhibit an extreme variation from sensor to sensor (across NORSAR variations of 5:1 in amplitudes are common) estimates of t* from common sources at the various sensors have a standard deviation of the order of .06 sec if one considers the .5-4 Hz band at the same array. (Slides 2 and 3) It would not be wise, therefore, to rely on absolute P wave amplitude variations to estimate relative Q. The amplitude variations in short period P waves have been shown to be the results of focusing on lateral heterogeneities in the crust and upper mantle under the array and is thus not related to Q. (Slide 4) This reasoning applies to all measurements of Q at individual recording sites using wave amplitudes. Only averages of amplitudes from many azimuths may be meaningful.

Another natural consequence of small scale lateral heterogeneities in the crust is that the seismic rays are diverted near the source to constitute large scale focusing-defocusing patterns at teleseismic distances (Hadley 1979, Butler and Ruff 1980). An example of such patterns after Hadley (1979) is shown in Slide 5.

Similarly the possible range of Q and its frequency dependence may be delimited by the range of observable signal frequencies and by the shapes of spectral ratios if frequencies up to 10 Hz are used. It can be shown, for instance that the spectra and waveforms of short period S waves put some quite tight upper limits on t^*_s along some paths, even assuming a delta function source. The shapes of wide band spectra in the short period band do not seem to support any rapid variation of Q with frequency, but frequency dependence of Q is necessary to reconcile long period Q estimates, if they are to be believed, with those obtained in the short period band. (Slide 6a, b)

In general studies of Q in narrower bands do not support rapid variations of Q with frequency (Der et al 1980, Lay and Helmberger 1981. Shore 1983). The arguments for frequency dependence of Q are more convincing, however, if one desires to reconcile long and short period Q estimates, but even in this case Q is very poorly constrained in the long period band for specific regions. Estimates from free oscillation may also be seriously biased relative to the real Q structures in the most typical regions of the world because of severe lateral variations in Q and velocities. Therefore the question of frequency dependence of Q can not be decided until more research is done, but it seems certain that any changes of Q with frequency is quite gradual along most paths.

In view of the absence of indications of strong frequency dependence I find the attempts of Lundquist and Samowitz (1982) to fit detailed pairs of absorption band models to interstation P wave spectral ratios quite premature, and only careful, broad band studies of Q extending to low frequencies will be able to determine the details of frequency dependence, instead of the narrow band data used by Lundquist and Samowitz (typically .5-4 z.).

If most of the anelastic losses occur in the upper mantle t* should not be strongly dependent on distance along paths in laterally homogeneous structures. This appears to be the case as long as we attempt to keep the structure separated. Shore's (1983) results and our own show the weak dependence of t* on distance. The somewhat higher values for these t* for shields are due to the fact that they were tied to PP measurements which because of the imperfect surface reflection in the short period band are likely to give upper limits.

For nuclear explosions the presence of surface refections and possible spall arrivals are a little more than a slight perturbation in the estimated Q value as long as one assumes a rather weak scaling effect on the spectra and a wide band of frequencies are considered, and no matter what the pP parameters are, the low and high t* paths yield spectra that are quite distinguishable.

In view of the extreme site variations in the short period band it is obvious that absolute amplitude modelling of waveforms is much less accurate for modelling Q. Likewise, the determination of other source parameters using absolute waveforms has been found quite nonunique Cormier (1982).

Although many papers in the literature that use $t_p^* = 1.0$ routinely in the short period band, regardless of the tectonic setting of the path, such results are usually quite poorly constrained (Der and McElfresh 1980). Moreover a lower t* would usually fit the data as well. In most cases t* trades off with source parameters, In one such study (Burdick and Helmberger 1977) the derived overshoot ratios were appropriate to a source medium of the consistency of rubber.

Exclusively high values ($t_p^* = 1.15$) by Helmberger and Handley would lead to unreasonably high bias estimates.

As a general comment I wish only to state that near and above 1 Hz the value of $t_p^* = 1$ appears to be quite inappropriate for most paths, not only shields. Results for shields give quite low absolute t* from PP (Shore 1983) or t_s^* from short period S phases (Der et al 1982). Similar results were obtained by Stewart (1983) from rise times of P waves ($t_p^* < .5$ sec).

In the following discussion of the effect of mantle Q on yield estimation we shall make the following implicit assumptions.

a) The frequency dependence of Q is weak along the various paths. Therefore the Q related amplitude variations will be correlated with the variations of the shapes of P wave spectra.

b) We assume that using large number of relative spectral ratios among recordings at various stations the source effects will cancel.

c) That explosions in hard rock will have source spectra adequately described by the von Seggern-Blandford model.

These assumptions agree with most observed facts we know of at the present time. Exceptions appear to occur for explosions in salt, clay, and, for high frequencies, at Yucca Flats. Within the above constraints we shall attempt to give our best estimates of the differences in the mantle attenuation under several major test sites and describe any data, less complete, for others. The estimates will be based on the best fits to available estimates of absolute and relative apparent t*.

After these preliminaries I shall review now the available evidence indicating major differences of mantle Q under some of the most important test sites.

The problem of primary concern in yield estimation is the so called "magnitude bias", i.e., variations of the m_b-yield relationships for various test sites caused by differing amount of amplitude reduction of P waves under various test sites due to mantle attenuation.

Another problem of great interest in nuclear monitoring is that of discrimination. While for explosions of large yield the M_s-m_b discriminant is very effective, at lower yields the seismic surface waves may not be detectable. To discriminate at such yields the use of the relative spectral contents of explosions and earthquakes was proposed. Explosions, being sources of small dimensions, can be expected to generate low frequency waves less effectively than earthquakes. Unfortunately, the spectral shapes are also affected by Q and the *a priori* knowledge of Q along the path is necessary for effective discrimination.

The mantle Q along the path is also a major factor in the information content of teleseismic signals. The accuracy in determining the details of the source functions or the amplitudes and delay times of surface reflections may be significantly reduced if the available signal bandwidth is diminished by low Q along the path. This will, in turn, affect the reliability of any correction of the amplitude for the surface reflection and thus the estimated yield itself.

Seismological evidence that Q variations pose an important problem in yield estimation emerged fairly early in the studies of seismic waves from U. S. Nuclear explosions. Seismic waves from the nuclear explosions Gnome and Salmon showed a strange asymmetry in the amplitudes and frequency contents of the observed P waves (Slide 7) which was noted in the sixties but largely forgotten or ascribed to differences in crustal structures soon after. Many recent workers agree that the first arrivals must have travelled through the upper mantle beyond a certain distance from these events so that the reduction in the high frequency contents of P in the western United States but not in eastern North America reveals a difference in mantle attenuation. Plots of apparent t* values derived from the estimated source spectra of these events show a constant slope characteristic of attenuation.

P wave spectra of explosions at the Nevada Test Site were shown in early studies to be characterised by a general lack of high frequency content beyond 4-5 Hz at teleseismic distances overall, while most explosions in the shield areas of Eurasia and recorded in the shield, were shown to have significant high frequency energy in their P waves with 8-10 Hz energy still above the noise level.

The path related spectral anomalies in P waves were also associated with anomalies in the amplitudes of the observed teleseismic P waves across North America reported in several major studies. The results of Booth, Marshall and Young (1974) demonstrated these for P arrivals. An apparently conflicting conclusion was made by Butler and Ruff (1980) asserting that there are no m_b differentials between the m_b values observed in the EUS and WUS or the US as a whole. Proper consideration of the total pattern of magnitude anomalies across the U.S. (Douglas and Marshall 1983) however totally invalidates this claim. It so happens that Butler and Ruff considered the averages of the west coast-intermountain belt and the NEUS-EUS regions respectively in which the first part was always over represented in station coverage. I think nobody after considering Slide 8 which shows the m_b results of the merged LRSM-WWSSN stations, could rightly claim that no m_b anomalies exist across the United States. Studies of S wave arrivals showed similar regional patterns of amplitude and spectral anomalies but much accentuated in a manner suggestive of energy losses in shear deformation in the mantle. Attempts to model these amplitude and spectral anomalies with variations of crustal effects failed. The amplitude anomalies could not be explained by the effects of even extreme crustal models (Slide 9). Note in this figure that the average WUS - EUS m_b difference is 0.35 m_b and that RKON is in the middle of the EUS population - not the highest amplitude station by any means.

Another early indication that the m_b-yield scales depend on the source region comes from studies of the M_s-m_b relationships for explosions (Slide 10) which showed a too low m_b for NTS explosions (or alternately too high M_s). This can easily be explained by greater attenuation under the WUS of the P waves. Simulations confirmed the regression results (Slide 11) and showed that a formula relating m_b to t^* $\Delta mb = 1.35\ \Delta t^*$ is approximately valid. The formula was empirically derived by fitting regressions to WUS and EUS crustally corrected Δm_b values by Der et al., (1979).

Since in estimating yields we are more interested in the geophysical properties of a few test sites rather than generalities about regional variations, during the late seventies ARPA undertook a major effort to measure the size of the intersite "magnitude bias" between NTS and a hypothetical shield station. Since the Soviet test sites are not accessible to us the shield site was typified by the station RKON (Red Lake, Ontario) on the southern edge of the Canadian shield. The stations HNME and IFME in Maine were found to be similar in surface geology to the Kazakh test and were therefore selected as possible analogs. These sites were also occupied at various stages of the experiment. At NTS we had two stations on the Climax stock, one at the Gold Meadows stock, two at Pahute Mesa and four at Yucca Flats. Later more stations located at various nuclear test sites in the western United States were added. These occupied the sites of the Gasbuggy, Rio Blanco and the Faultless explosions. The site of another explosion in granite, Shoal, was occupied earlier by an LRSM station. (Slide 12) After approximate corrections for the crustal structure were applied the trace amplitudes revealed lower amplitudes at all WUS sites relative to RKON. (Slide 13) The m_b differences (log A/T) are reduced from these single (log A) differences. The measurements of apparent t* are especially revealing (Slide 14). There appears to be a significant difference in the frequency contents of all teleseismic arrivals at the WUS sites relative to RKON. The two sites in the NEUS, HNME and IFME, appear to have an intermediate position between the shield and the WUS. As we shall see the Kazakh t* values are low so that these intermediate t* values suggest that the NEUS stations HNME and IFME are not good analogs of the Kazakh test site. The differences in the frequency contents of P waves are also apparent in the average differentials in dominant period at stations with similar instrument responses.

Similar regional dependence in t* was also found for P wave arrivals from Soviet nuclear explosions across North America (Slide 15) by Shore (1983). For this data set the RKON-Basin and Range t* differential is greater than in the SDCS experiment.

This experiment has found that the test sites within the WUS fit well into the regional pattern of mantle attenuation in the same region.

The average t^* differential between RKON and the station at the Climax Stock at NTS was near .2 sec which corresponds to about a 0.27 magnitude differential assuming a constant Q.

The patterns of regional attenuation found in the NTS experiment were further confirmed by and fit well into the results of broader regional studies of mantle Q under the United States (Der et al., 1982, Lay and Helmberger 1980). The studies of short period S waves yielded especially dramatic examples that confirm the reality of lateral Q variations. There can be, therefore, hardly any doubt about the existence of lateral variations of Q under the United States. Slide 16 shows a good example of variations in the amplitudes and wave periods of short period S waves from deep earthquakes across the United States due to lateral changes in mantle Q. Burdick (1982) has laso concluded that most of the United States is underlain by one of two types of mantle.

Following the NTS experiment we have undertaken an extensive study of the spectra of teleseismic P waves from NTS and Kazakh nuclear explosions as recorded at a large number SRO, LRSM and array stations. We are now dealing with a different, more difficult, problem; determination of t* from the spectra by assuming the source spectrum instead of cancelling the source relative measurements.

The von Seggern-Blandford model was used as a preliminary source model. Despite the possiblity of multiple arrivals and other complications, comparison of the spectra from explosions with varying estimated yields seemed to obey the cube-root scaling law we have used for the von Seggern-Blandford model. Any uncertainties of the assumed source spectra will, in this case, affect the t* estimates. We have accumulated more than 100 spectra during this experiment. By comparing the spectra of NTS and Kazakh explosions for events with comparable yields the major difference in the spectral contents in P waves from the two test areas is quite obvious. (Slides 17a, b show examples from NORSAR). At arrays the overall spectral shapes remain consistent large despite site amplitude variations!

In order to interpret the total data set and to exploit any internal consistencies in the data we have assumed the simple statistical model;

$$t^*_{ij} = t^*_i + t^*_j + e_{ij}$$

Where t^*_{ij} is the measured average t* from test site i at station j, t^*_i and t^*_j are the test site and station terms and e_{ij} is an error term. In this model we ignore the distance dependence, which at teleseismic distance is small, and any dependence on azimuths. We therefore attempt to determine average upper mantle Q contributions to t* under the test sites and receivers averaged in the small cone in the mantle sampled by teleseismic ray paths.

Since the test site and station terms trade off against each other, no unique solution exists unless we impose some constraint. Since in the SDCS experiment we had a measured t* difference between Pahute Mesa and RKON of .2 sec we have adjusted the differences in the regression results to this independently determined value. (Actually this differential was measured more accurately than the accuracy of the regression results).

By looking at spectra of Yucca explosions at regional distances we have concluded that the von Seggern-Blandford granite RDP used in our analyses is not valid for these events. We decided therefore to use the Yucca data only for refining the station terms (which are determined by spectral ratios). Determination of Yucca RDP-s requires more work. Actually the direct Yucca t* estimates are not needed to determine Yucca test site t* because we have a very large data set of reciprocal measurements that show that Yucca Flats are not significantly different from Pahute Mesa with regard to mantle Q. The regression on the other hand shows a rather large Δt^*, between Pahute Mesa and the Kazakh test sites. (Slide 18)

The station terms on the left do not vary much, although WHYK and ZOBO are high. NORSAR and NPNT are low. The test site terms show a great difference between NTS and Kazakh (Degelen and Shagan). Based on these results one can say with certainty that NTS and Kazakh must have quite sizeable differences in the attenuative properties of the upper mantle, but there may be a sizeable uncertainty in the NTS measurements due to the narrow band P wave spectra from NTS. The station terms are less affected by this. Therefore we cannot base our estimate of the bias between the two test sites on the results of these direct, not relative, t^* estimates alone.

The regression gives a higher Δt^* bias for Pahute Mesa relative to Kazakh than the OB2NV-RKON difference. Can it be that Kazakh and NORSAR have higher Q in the underlying mantle than RKON? This appears to be the case if one compares the spectra of P waves from common events at RKON and NORSAR. The corresponding differential in t^* is about .1 sec. Therefore it is possible to substantiate a NTS-Kazakh bias of the order of .3 in t^*. In the course of making t^* estimates from Pahute Mesa NTS explosions we did not note any major deviation in the source spectral shape from the von Seggern-Blandford model, and thus the .3 bias seems to be reasonable considering all the evidence. Another relevant question; can it be that the Kazakh explosions put out anomalously high amounts of high frequency energy? This however, can be ruled out outright by considering the extremely high frequencies in the P wave spectra up to 10 Hz that would be impossible to observe with substantially higher t^* even if one assumed a flat source spectrum. The t^* estimates from Soviet PNE-s are essentially similar to those obtained from Kazakh explosions.

An equivalent of the reciprocal experiment performed at NTS, although not as exhaustive, was made possible by published data from a Soviet CISS station near Semipalatinsk. Comparison of the spectra of Alaskan earthquakes at NTS and at this station resulted in a bias estimate in t^* of .24 sec by Murphy and Tzeng (1982).

Blandford (1981) comparing carefully selected Soviet explosions to Piledriver at the common station NPNT obtained a Δt^* of 0.25 as a best estimate.

Burdick (1982) comparing the WUS and the Eurasian shield via S-SS long period delays concludes that the t^* difference between these broad areas should be 0.2.

Considering the total evidence it can be said that the NTS-Kazakh t^* bias due to excessive attenuation under the Basin & Range must be in the range of .2-.3 sec. It is probable that it does not exceed .3 but is almost certain to be more than .2. Assuming the formula $\Delta m_b = 1.35 \Delta t^*$ the magnitude difference corresponding to this t^* difference would be 0.27 to 0.40 in m_b. However, in the SDCS experiment we noted that the m_b differential between NTS and RKON was 0.1 less than that implied by the Δt^* values. This was attributed to focusing under NTS. This correction must also be applied in the present study so that our final bottom line is a Δm_b in the range 0.17 to 0.30 m_b.

STUDIES OF P WAVES FROM OTHER TEST SITES

Algeria

A site of considerable interest in yield determination is the Hoggar massiv in Algeria where some of the French nuclear tests took place. This site is presumably located over a so called "hot spot" and is characterized by incipient rifting, recent volcanism and thinning of the crust. Comparison of P wave spectra from the nuclear explosion Rubis at Hoggar and a Kazakh explosion at the AWRE array EKA indeed shows a decrease in the high frequency content of P waves from Hoggar compared to those from Kazakh. This supports the idea that Hoggar is indeed located over a "hot spot". Observations of P waves spectra of nuclear explosions at Hoggar also show the same decrease of high frequency content in P waves relative to shield type of paths.

Comparing the apparent t* of the nuclear tests at Hoggar to those at NTS, the Hoggar site appears to be intermediate between the shield and NTS with respect to attenuation in the underlying mantle.

Novaya Zemlya

Novaya Zemlya nuclear explosions observed at high Q stations yield low t* estimates indicating that the Q in the mantle under this source region is high, and comparable to that under Kazakh.

Lake Baykal Area

Seismic P waves recorded at LASA and NORSAR for two nuclear explosions in the Baykal area did not show any indication of encountering a low Q upper mantle along these paths. (Sobel et al., 1977). In a more detailed study covering a wider range of azimuths Savino et al., (1975) have found that the spectral discriminant failed along some azimuths (Slide 19). This can be interpreted as the effect of low mantle Q under the rift zone, also indicated by other geophysical studies.

Indian Nuclear Explosion

The rise times of P waves and spectral analyses of P from the Indian underground explosion at NORSAR are consistent with a high Q path.

Soviet PNE-s

As we have already mentioned Soviet PNE-s must be located in the areas of high mantle Q considering the rise times of P and their spectra. This includes known shots in salt.

French Nuclear Explosions in the Pacific

These have not been studied much with respect to Q but published P wave spectra at the Hagfors array in Sweden (Nedgard, 1978) are indicative of high mantle Q under these test sites.

Amchitka Tests

Analyses of the U. S. nuclear tests on Amchitka island usually result in t^* at stations in the contiguous U. S. of the order of .4-.5. Considering the fact that the mantle Q under some U. S. stations is low, this indicated low or moderate attenuation in the source area. These admittedly scarce results agree with the findings of Barazangi et al., (1975) that there is no extensive back-arc attenuation zone in the mantle near the Aleutian islands.

CONCLUSIONS

Spectral analysis of P waves crossing the mantle under various test sites has revealed considerable variations in the attenuative properties of the mantle which must be considered when computing yields of nuclear explosions. The t* bias between the NTS and Kazakh test sites must be in the range 0.2 to .3 sec. There are indications of anomalous zones of attenuation under the French test sites in the Sahara and near some suspected nuclear explosions in the Baykal area. ~~The~~ mantle attenuation under the Sahara test sites appear~~s~~ to be intermediate between a shield and the western United States. ~~The~~ other test sites for which we have ~~any~~ relevant results seem to be in areas of low mantle attenuation. ~~area.~~ ~~The~~ patterns of attenuation found by ~~the~~ analyses of nuclear explosion data fit roughly in the worldwide patterns found in more general studies of plate tectonics. No~~ne of the~~ test sites are located in areas of extremely low Q zones in the mantle, such as those found under some back-arc basins.

Empirical analyses and synthetic simulations give the formula:

delta m(sub b) = approx. 1.35 delta t*(sub p)

$$\Delta m_b \approx 1.35 \quad \Delta t^*_p$$

for deriving the m_b (sub b) bias from a t* differential.

~~The~~ regional patterns of lateral Q variations found by spectral analyses are ~~also~~ well correlated with several ~~a number of~~ other relevant geophysical parameters, such as P_n (sub n) velocities, travel time variations of P and S waves, heat flow, and upper mantle conductivity. Although care must be exercised in the use of such quantities for deriving mantle Q since these other parameters are not directly related to Q, they can be used as diagnostics in the absence of direct data.

Indirect methods of estimating the mantle attenuation using body wave (P_n, S and SS velocities have been proposed (Marshall, Springer and Rodean 1979, Burdick et al., 1982) and may be of some value in inaccessible regions. (Slides 20 and 21). Strictly speaking, however, body wave velocities may only be loosely related to Q and there are indications that body wave travel times worldwide may not correlate with

Q and magnitude variations as well as in the U.S. Nevertheless the promising results of such techniques merit further study.

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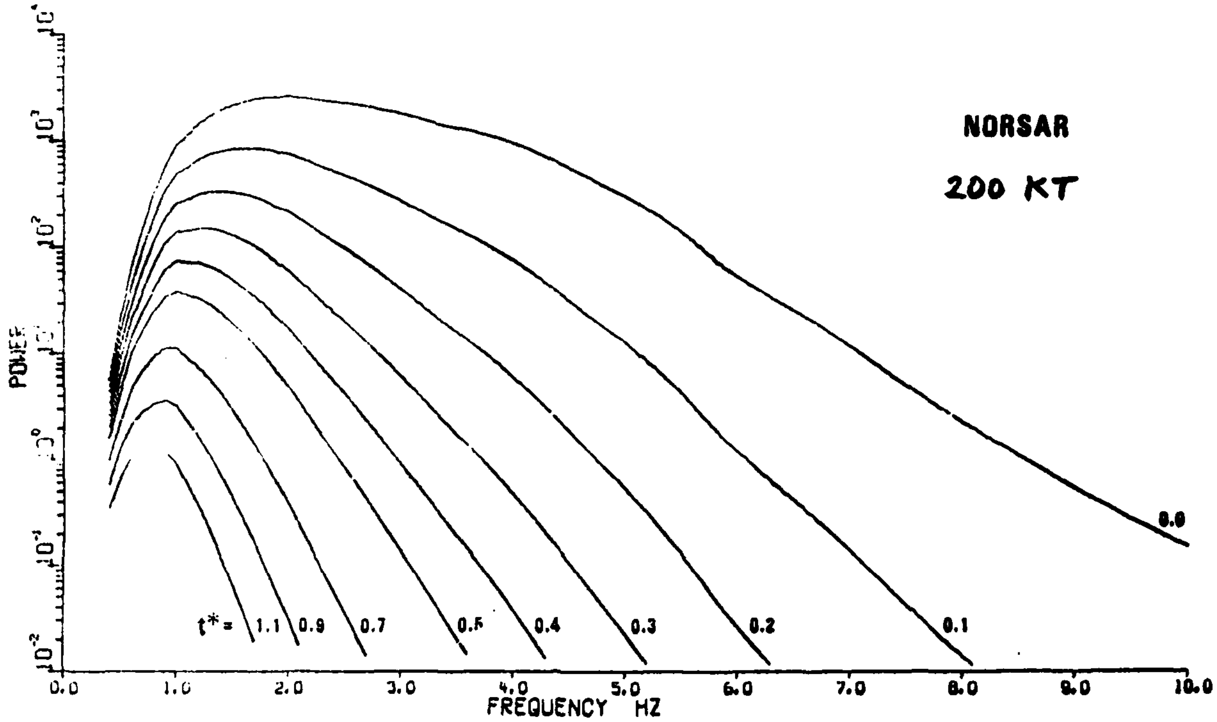
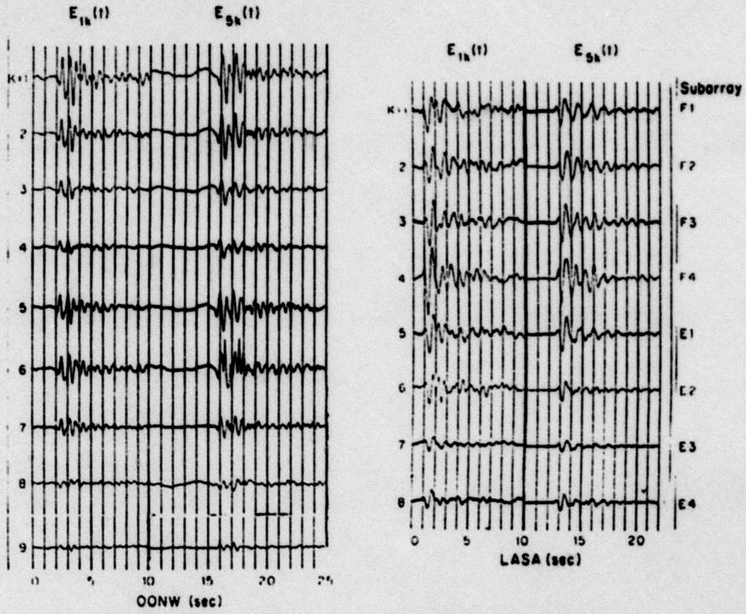
LIST OF SLIDES

Slide	Title
1	Variations in the shapes of the P wave spectra from a 200 kt nuclear explosion with t*. At low t* these shapes are very diagnostic, but the accuracy of estimation decreases with increasing t*.
2	Illustration of the variations of P wave amplitudes across two arrays.
3	Scatter in apparent t* across NORSAR. While the P wave amplitude commonly vary by a factor of five, the apparent t* have a low (.06 sec) standard deviation.
4	Azimuth-distance dependent P wave focusing patterns at LASA. After Chang and von Seggern (1981).
5	Azimuthal variations in amplitudes for various NTS events indicating near source focusing effects (After Hadley 1979).
6a,b	Some t*-vs-frequency curves (a) and the associated spectral ratio shapes (b). Small changes results in drastic and detectable changes in the spectral ratios putting severe limits on acceptable t*-vs-frequency relationships. In genera; rapid changes can be ruled out.
7	P waveforms for the Salmon nuclear explosion. Variations in the frequency content at upper mantle distances reflect lateral variations in Q.
8	Station terms for M_b for the merged LRSM-WWSSN networks across the United States (Douglas and Marshall 1983).
9	Regression analyses of the m_b anomalies vs station crustal amplification factors (After Der, McElfresh and Mrazek 1979). This demonstrates that crustal amplification cannot explain the anomalies.
10	M_s-m_b relationships for NTS and Kazakh at AWRE stations. (After Marshall and Basham 1972).
11	Results of some theoretical simulations for deriving a t* vs m_b dependence. The m_b = 1.35 t* line is superposed.
12	Map of stations used in the "NTS experiment". (NTSE)
13	Averaged relative trace amplitudes derived form the NTSE.
14	Relative t* results from the NTSE.


NORSAR
200 KT
POWER
FREQUENCY HZ
t* = 1.1
0.9
0.7
0.5
0.4
0.3
0.2
0.1
0.0
0.0
1.0
2.0
3.0
4.0
5.0
6.0
7.0
8.0
9.0
10.0

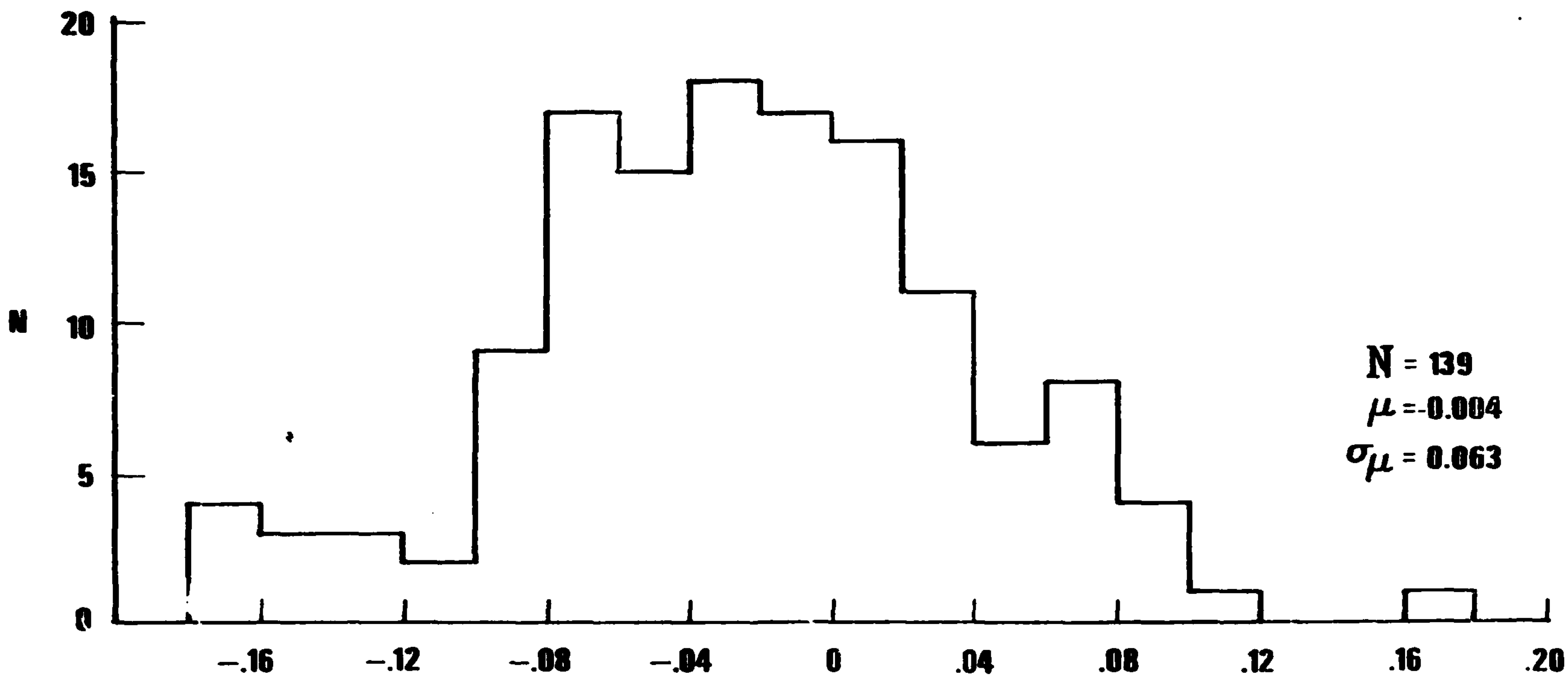
Slide 1.



Seismograms of events 1 and 5 recorded at the sites of OONW and Lasa, the smallest and largest arrays in this study.

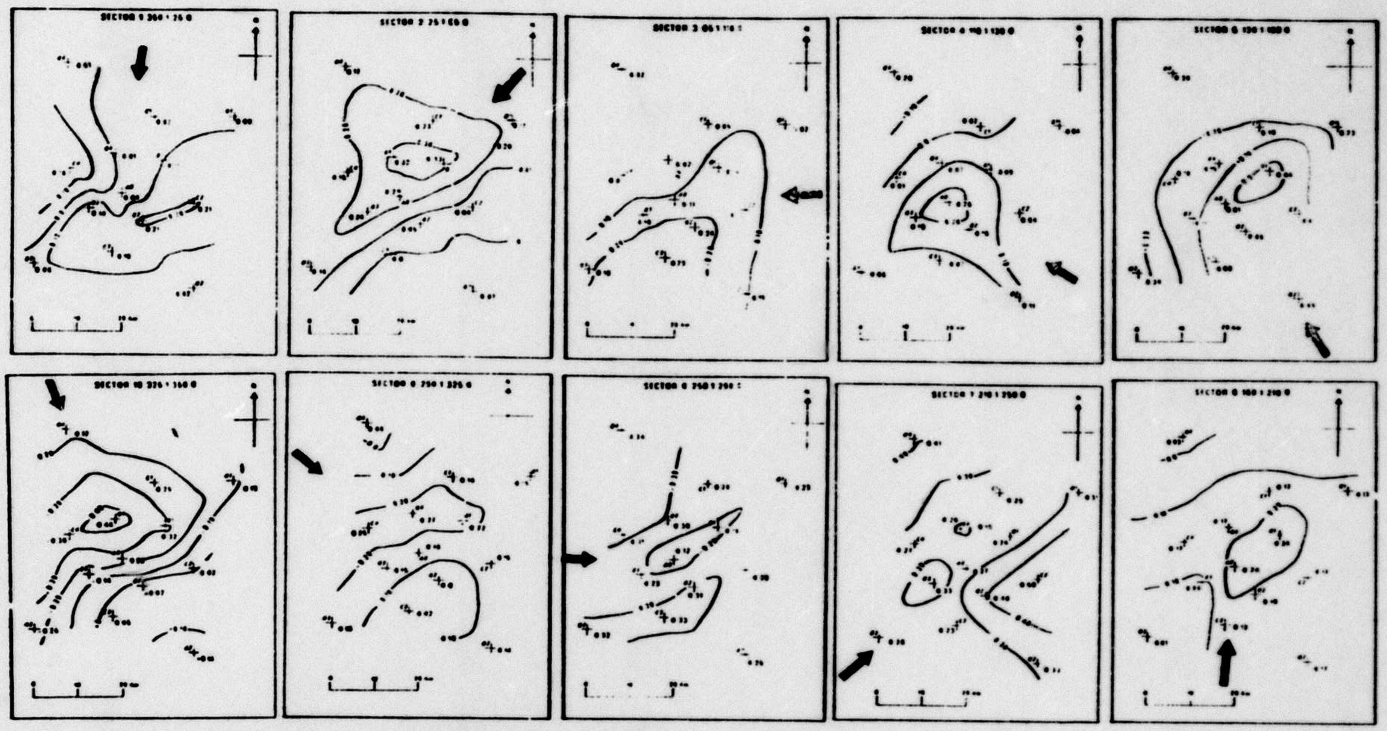
Slide 2.

RELATIVE t* MEASUREMENTS ACROSS NORSAR



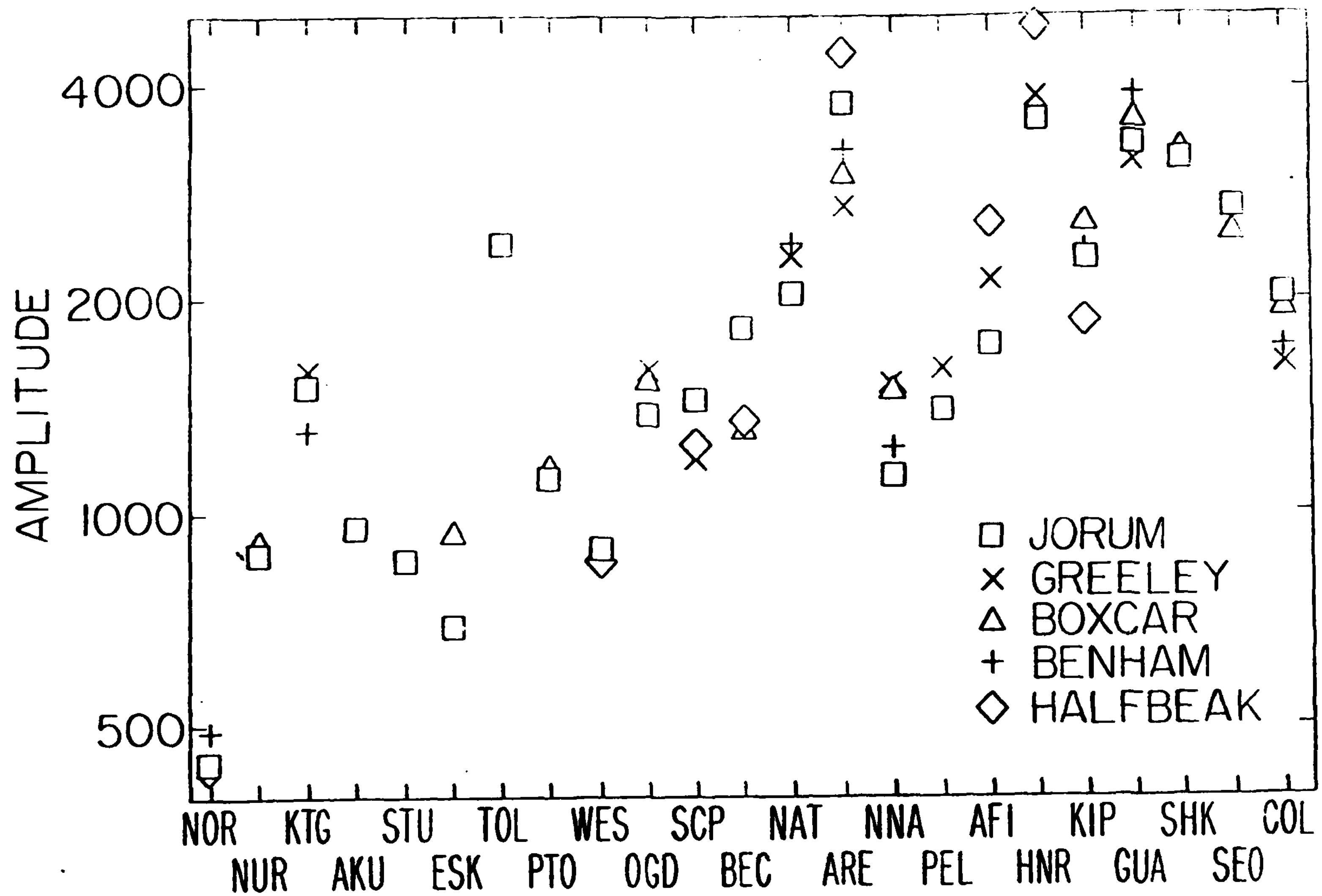
Relative t* between subarrays at NORSAR for ten teleseismic events. The histogram shows that the scatter of Δt^* compared to that of amplitudes is small (σ = 0.06 sec). This illustrates the relative stability of spectral measurements.

Slide 3.



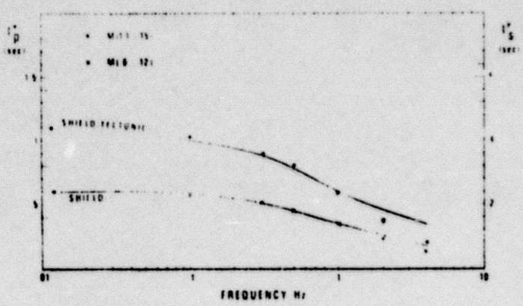
Contour maps of average m_b bias in 10 divided sectors. The variations in contour lines suggests the influence of variations in local crust-upper mantle structure.

Slide 4.



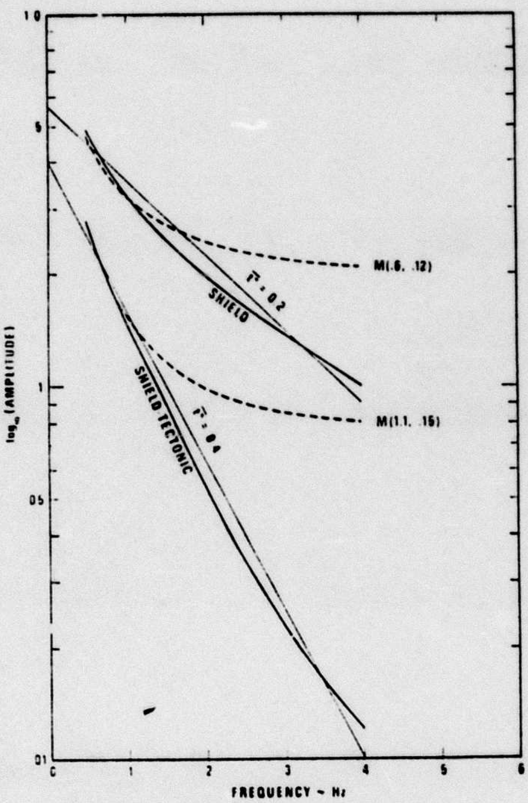
Azimuthal variations in amplitudes as observed for five tests located throughout Pahute Mesa (NOR = 10°, COL = 336°). The absolute amplitudes for each event have been adjusted in order to minimize scatter that results from variations in source strength.

Slide 5.



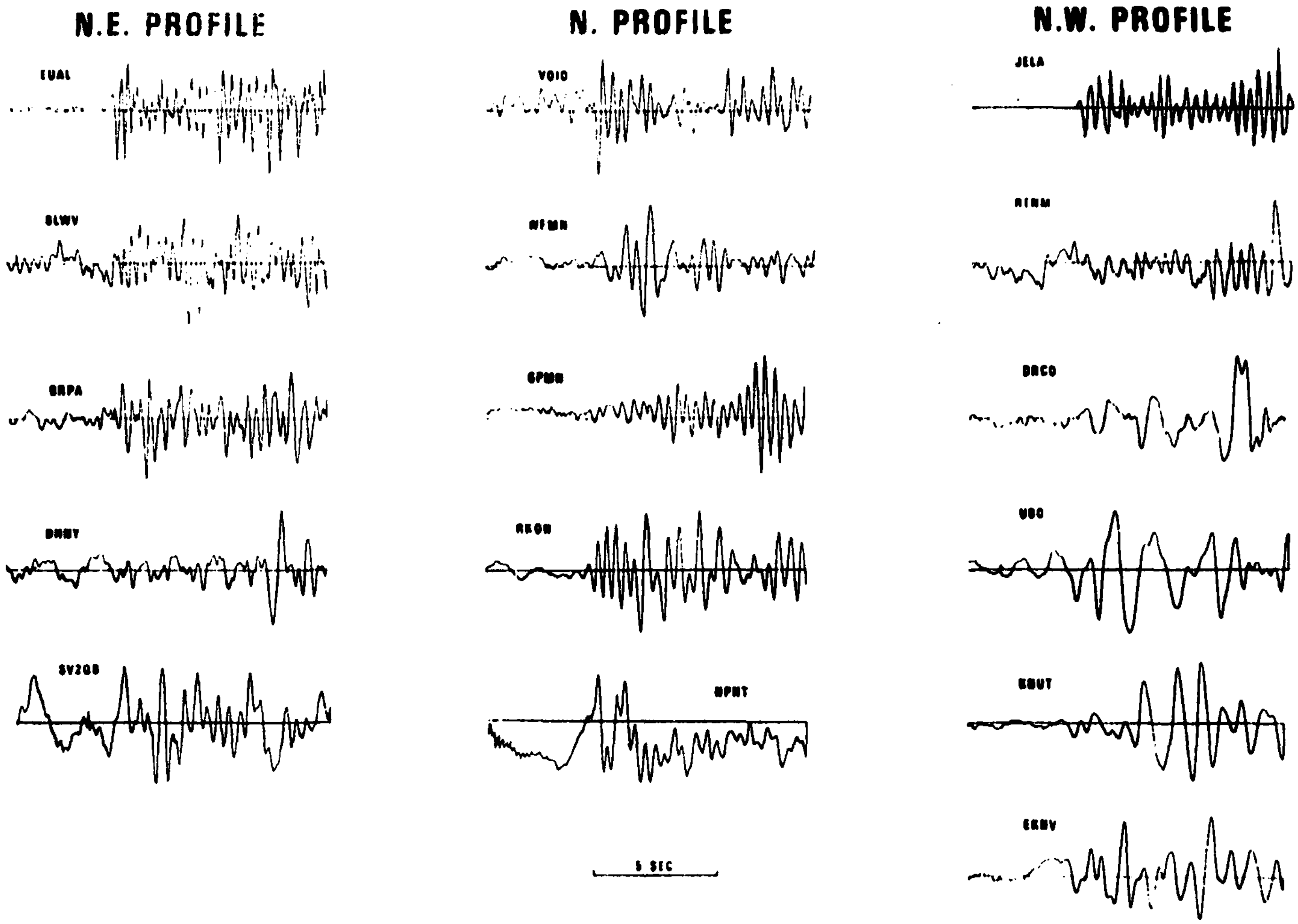
Proposed t_p^* versus frequency curves for shield and mixed shield-tectonic paths. The dots and crosses denote the closest fitting single absorption band models of Minster (1978). The notation $M(a, b)$ denotes a single absorption band model with a long-period asymptotic value of a and a τ_m value of b.

Slide 6a.



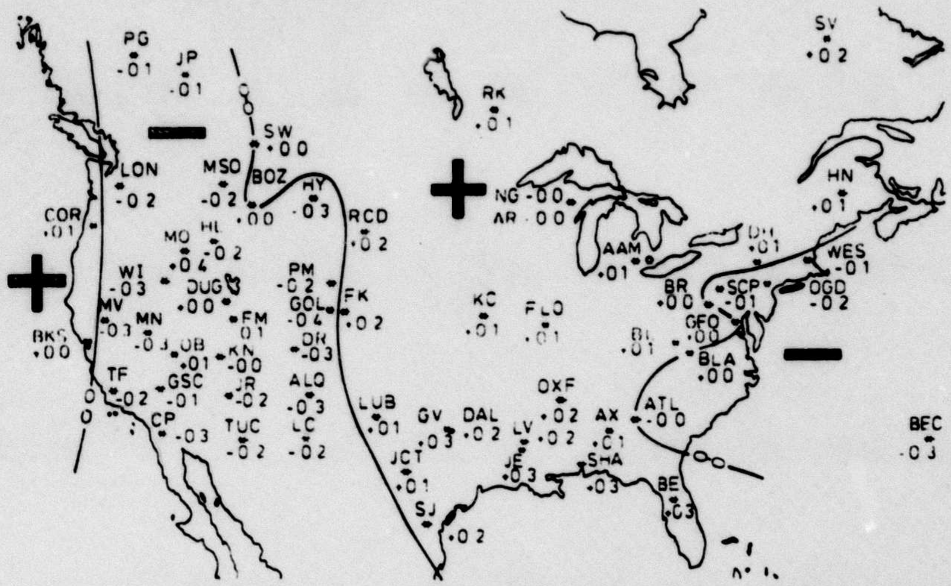
Observed-to-source spectral amplitude ratio shapes implied by the proposed t_p^* versus frequency curves and the absorption band models in Fig. 5a.

Slide 6b.



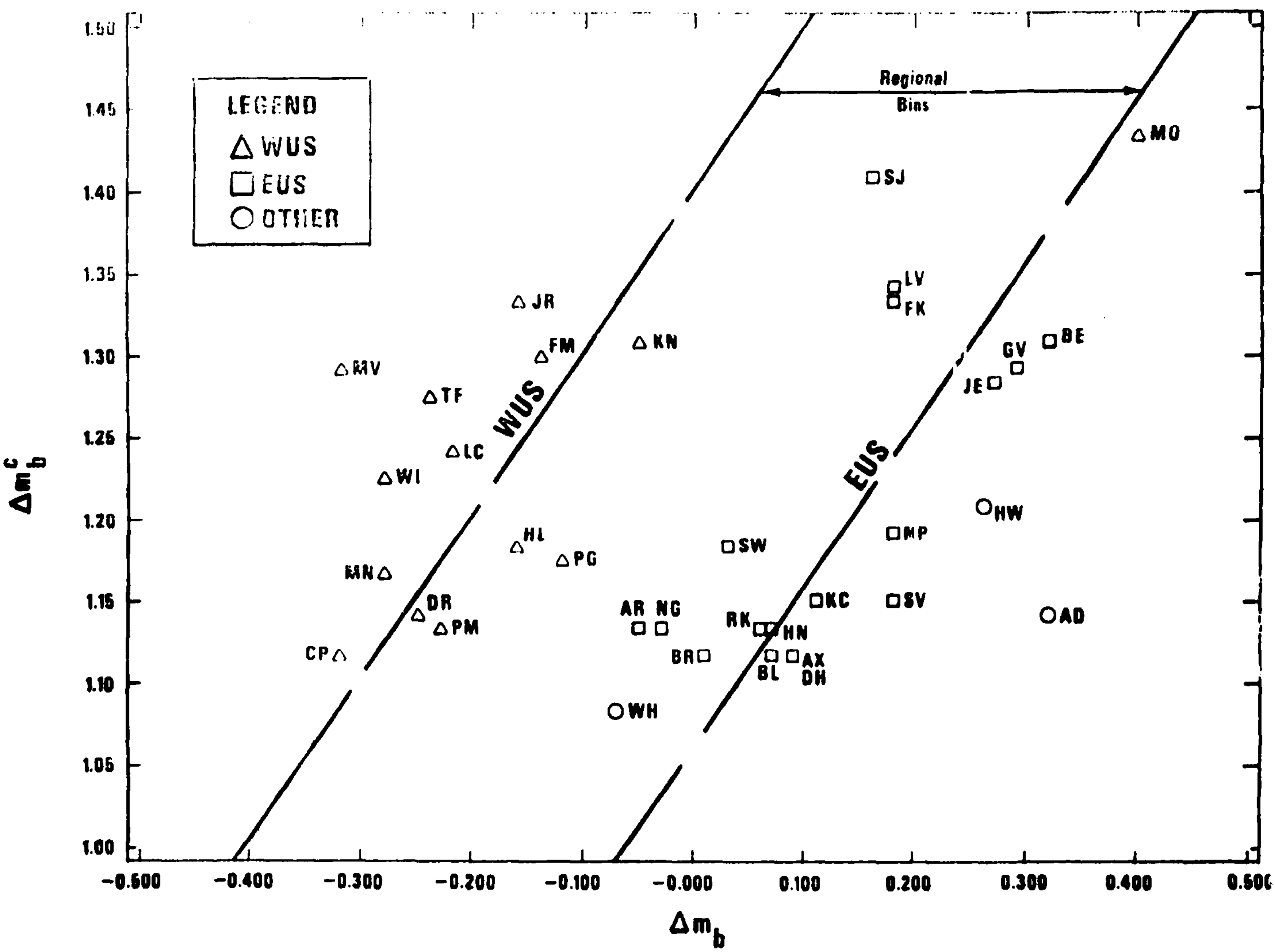
P-wave seismograms used.

Slide 7.



Station effects obtained from analysis of WWSSN observations together with effects estimated by Booth *et al.* (1974) for the LRSM stations. Contours divide areas of generally positive effects from generally negative effects.

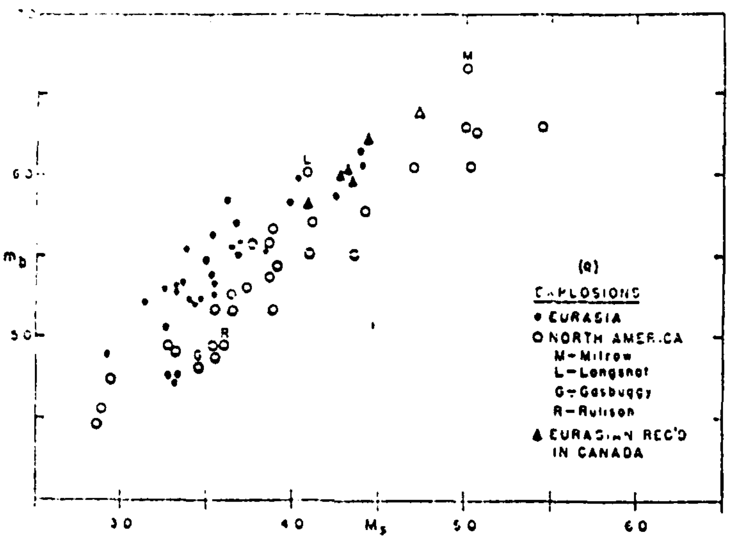
Slide 8.



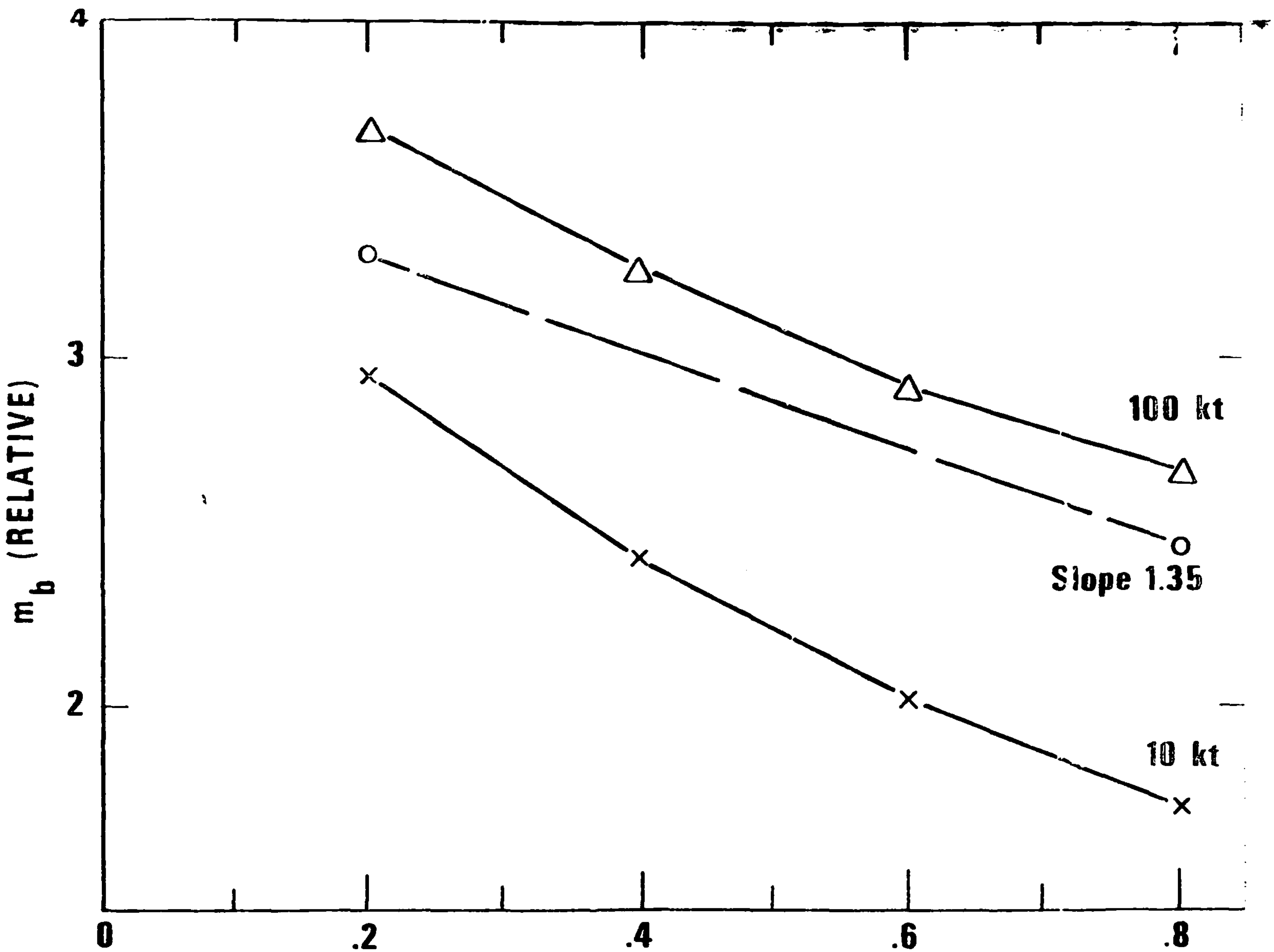
Magnitude residuals for Booth et al (1974) plotted against the logarithms of crustal amplification factor A_c. The data points tend to cluster around two regression lines, one for the EUS, the other for the WUS.

Slide 9.

Discrimination between earthquakes and underground explosions

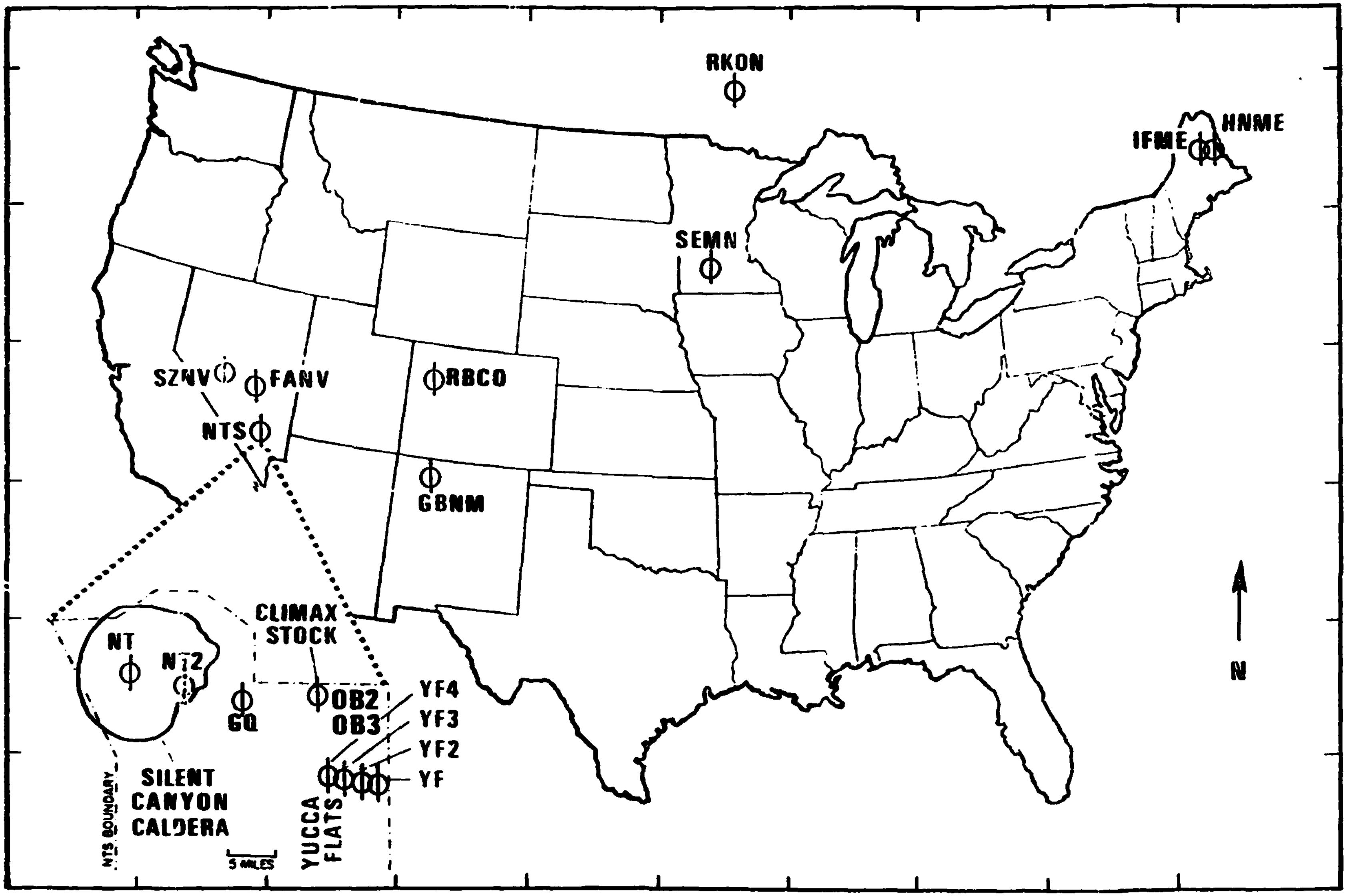


Slide 10.



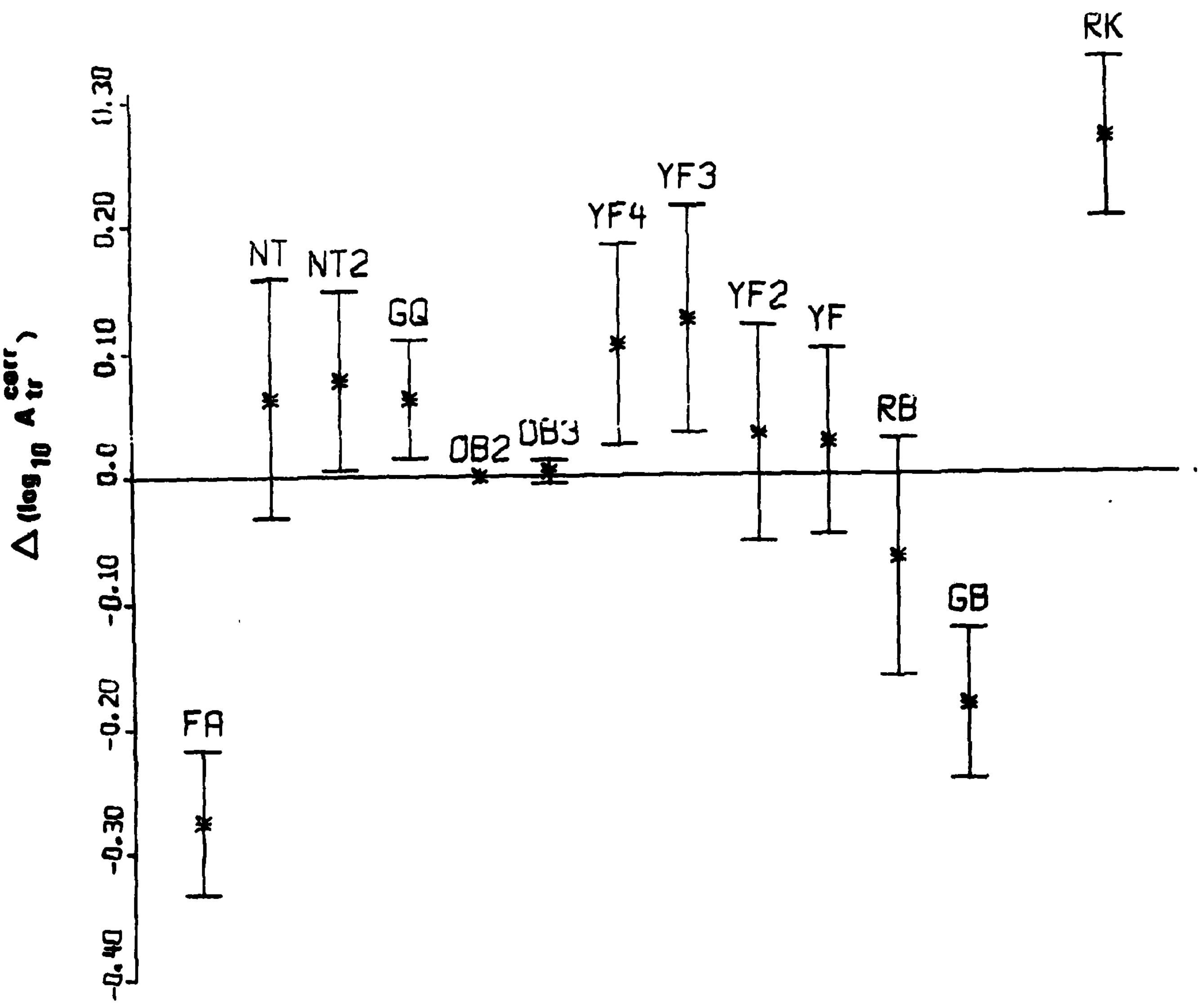
Decrease of m_b computed from synthetic pulses for a 10 kt and a 100 kt nuclear explosion due to increasing t^*. The empirical relationship $\Delta m_b \sim 1.35\ \Delta t^*$ is sketched in for comparison. The overall slopes of the three curves are similar.

Slide 11.



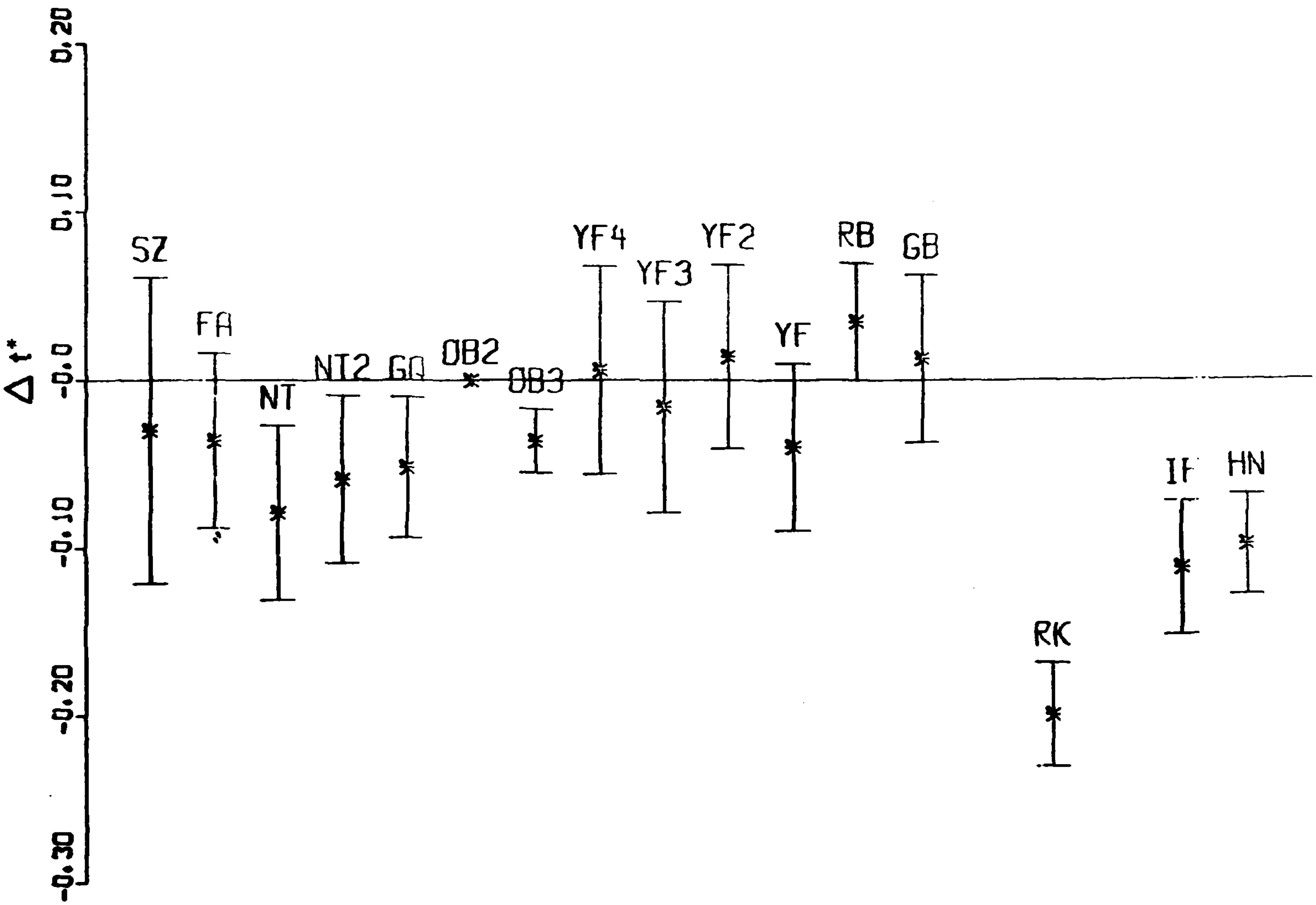
Locations of the SDCS and LRSM stations analyzed in detail in Part I of this report.

Slide 12.



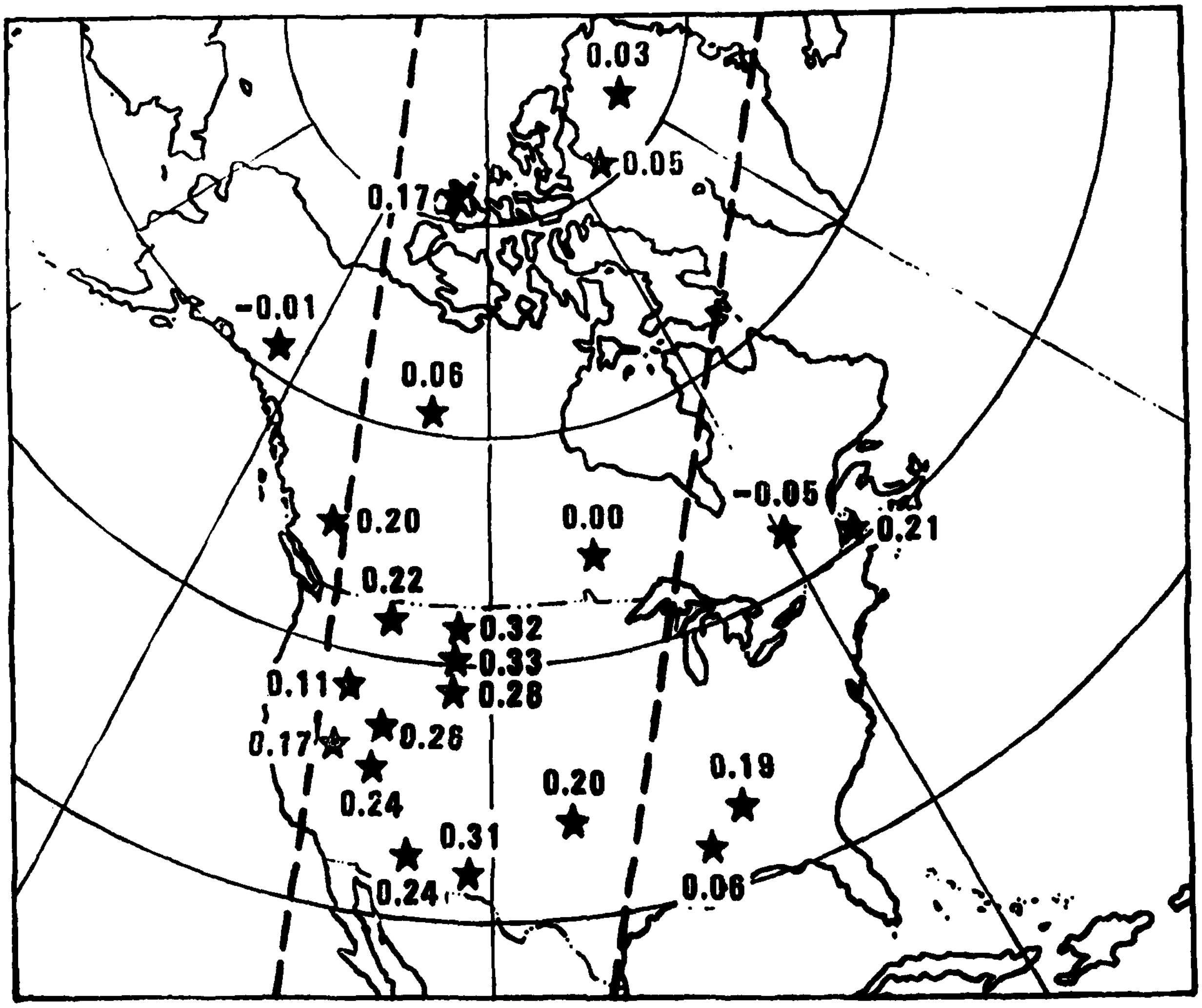
Trace amplitude levels relative to OB2NV ($\log_{10}$ units) with crustal corrections applied.

Slide 13.



Δt^* values of selected SDCS and LRSM stations relative to that of OB2NV. RKON on a shield has the lowest t^*, while the WUS stations have the largest. 95% confidence limits are indicated by bars.

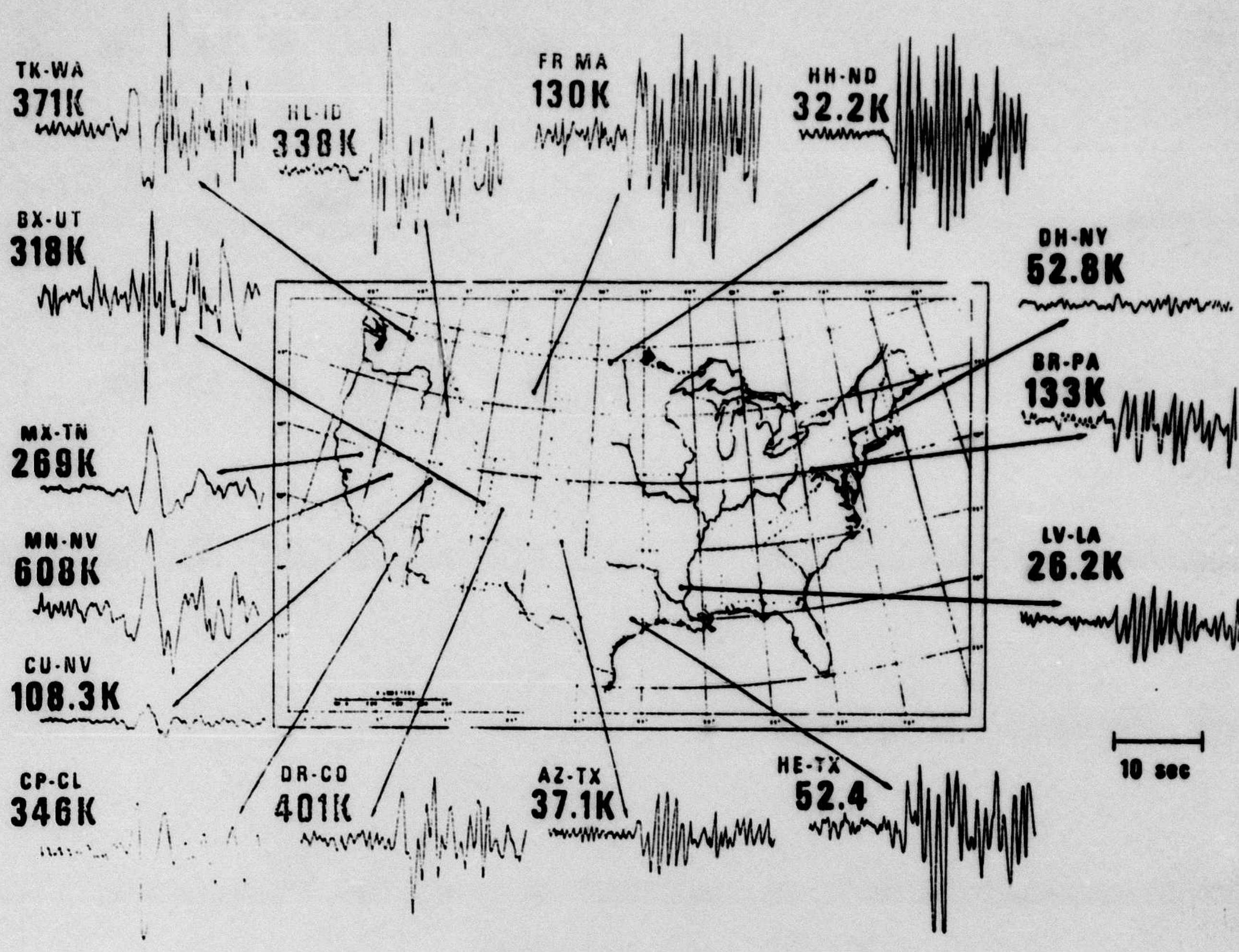
Slide 14.



t* station corrections.

Slide 15.

PERU-BRAZIL BORDER 09.25 71.5°W
10 NOV 63 OT = 01:00:38.8



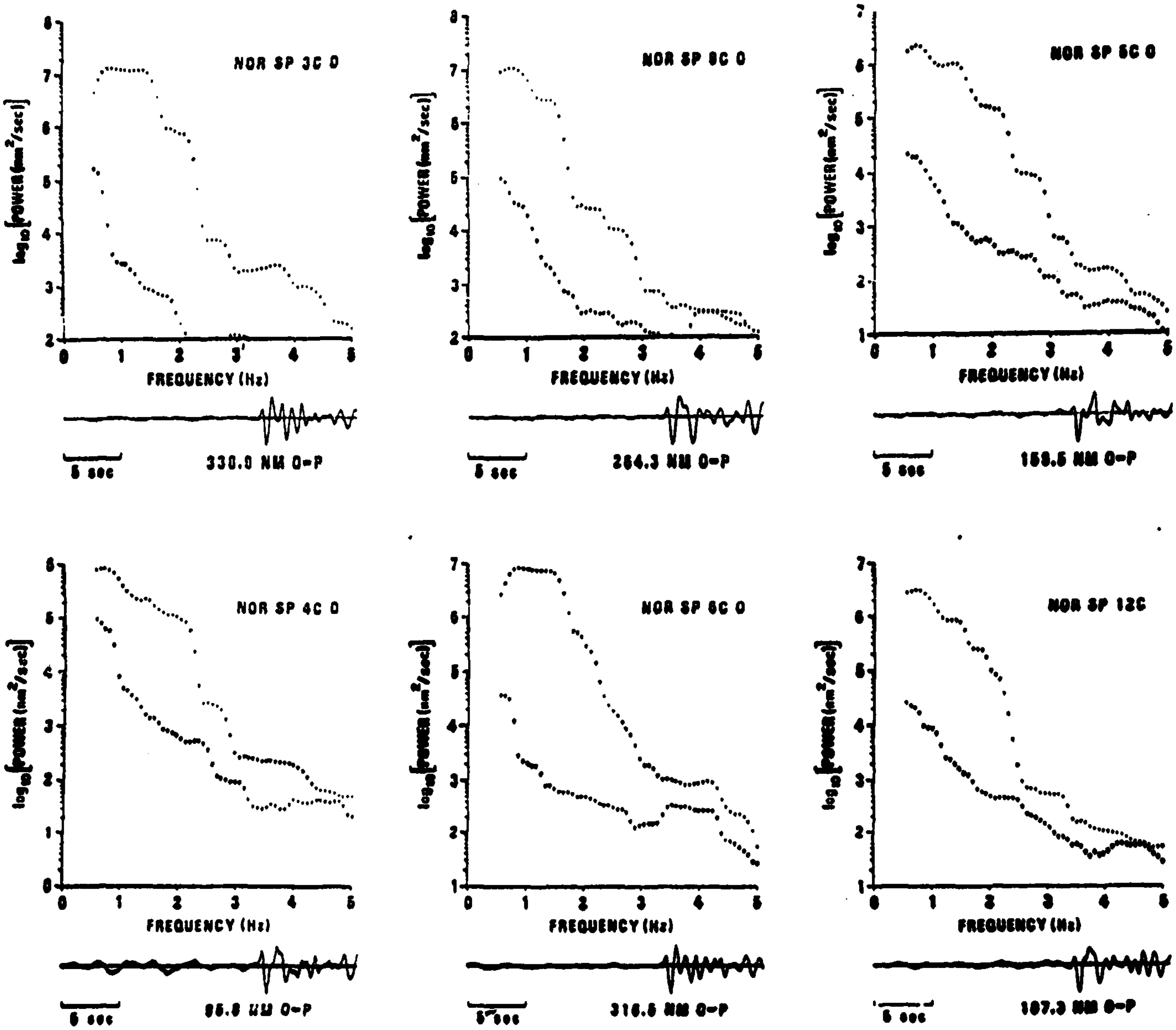
Short period SH

Tracings of short-period SH phases at LRSM stations across the United States. Depending on the frequency content of the signals, the time domain manifestations of anelastic attenuation vary, but the overwhelming majority of the signals show a diminution of amplitudes and the decrease of high frequency content in most of the WUS with especially severe effects in the southwestern United States. No corrections for radiation patterns were made in these figures. Instrument gains are shown on each trace.

Slide 16.

28 OCT 75

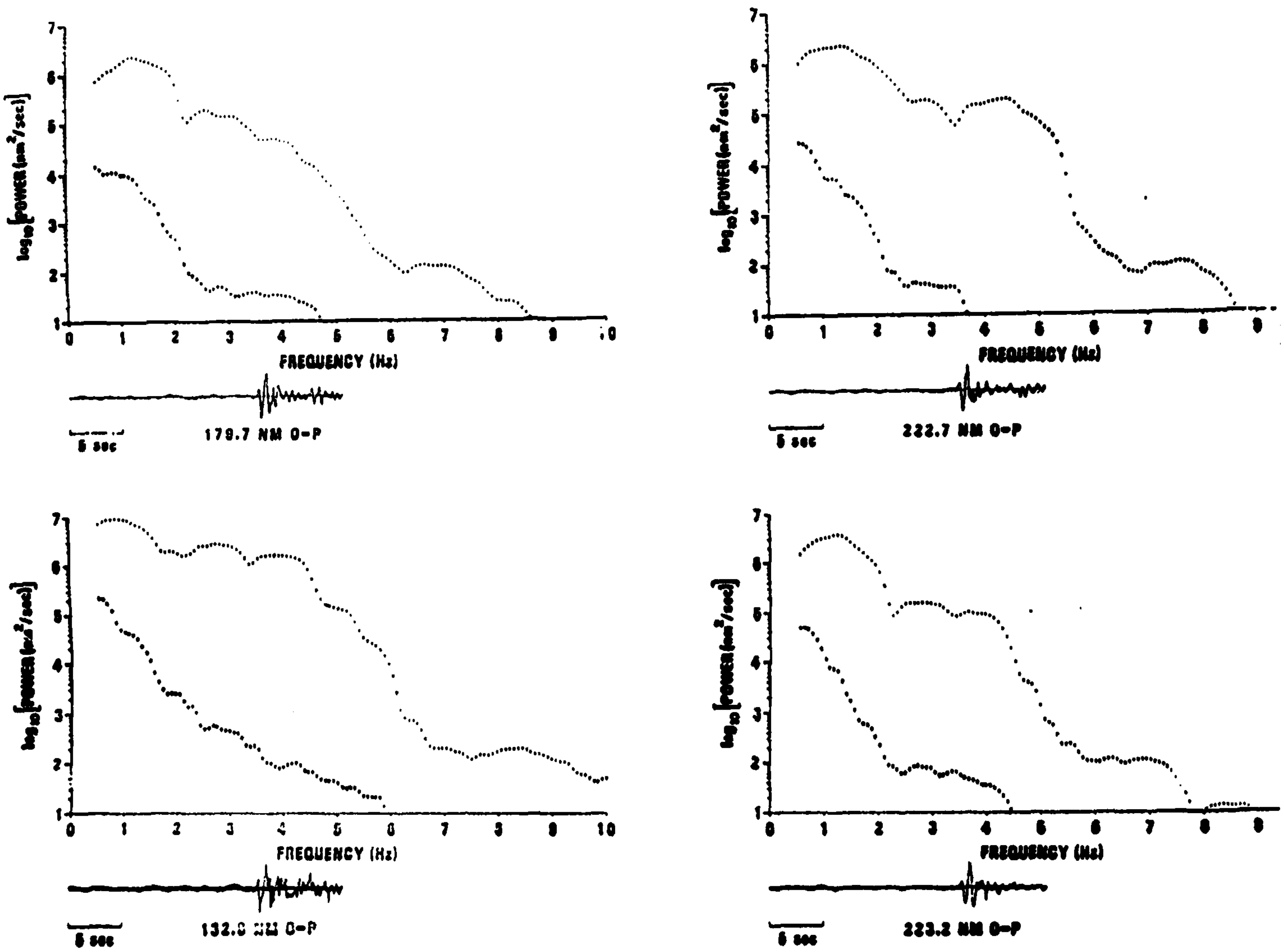
KASSERI AT NORSAR



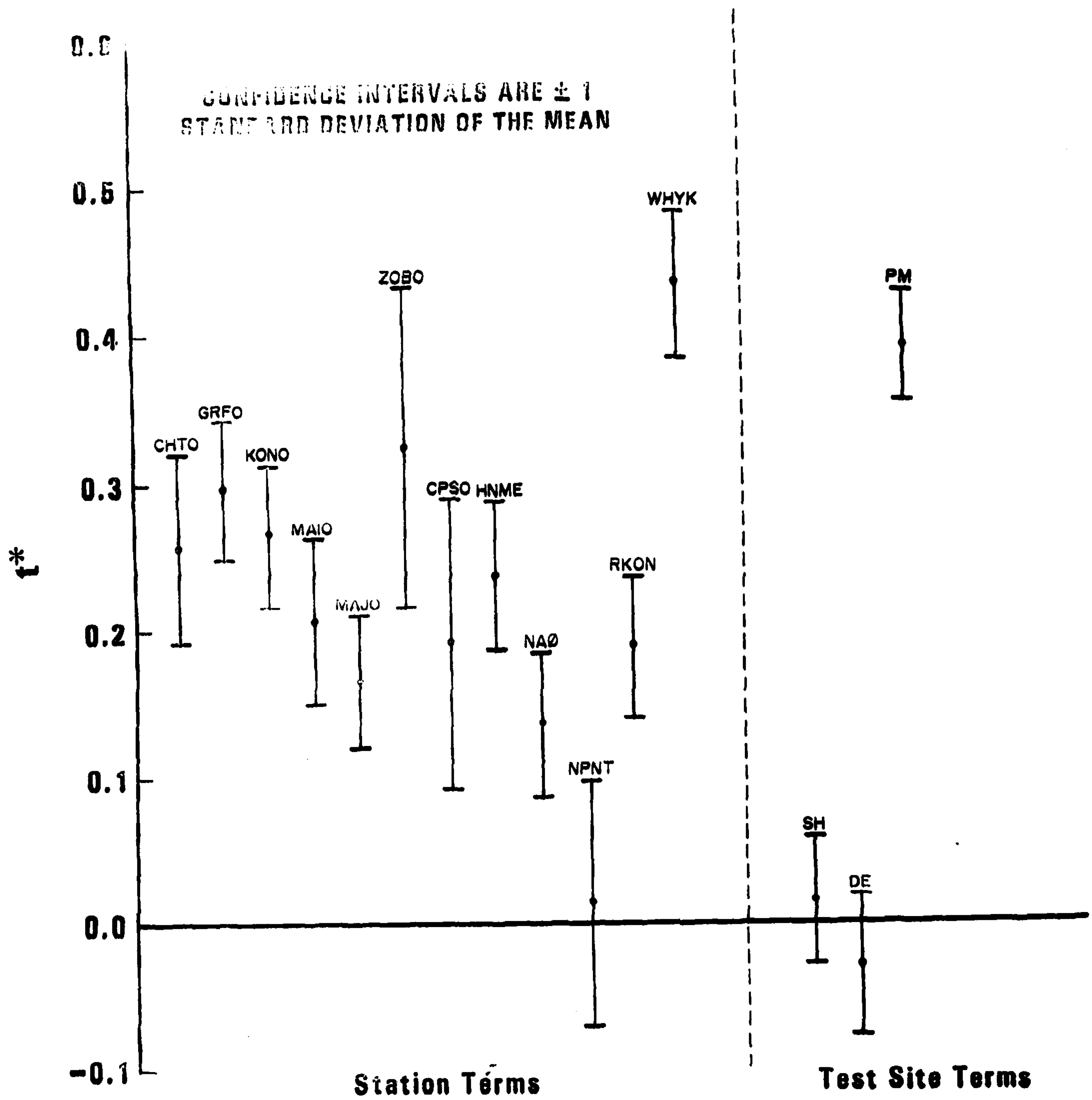
Slide 17a.

10 FEB 72

KAZAKH SHOT AT NORSAR



Slide 17b.



~~Alternate~~ Slide 18.

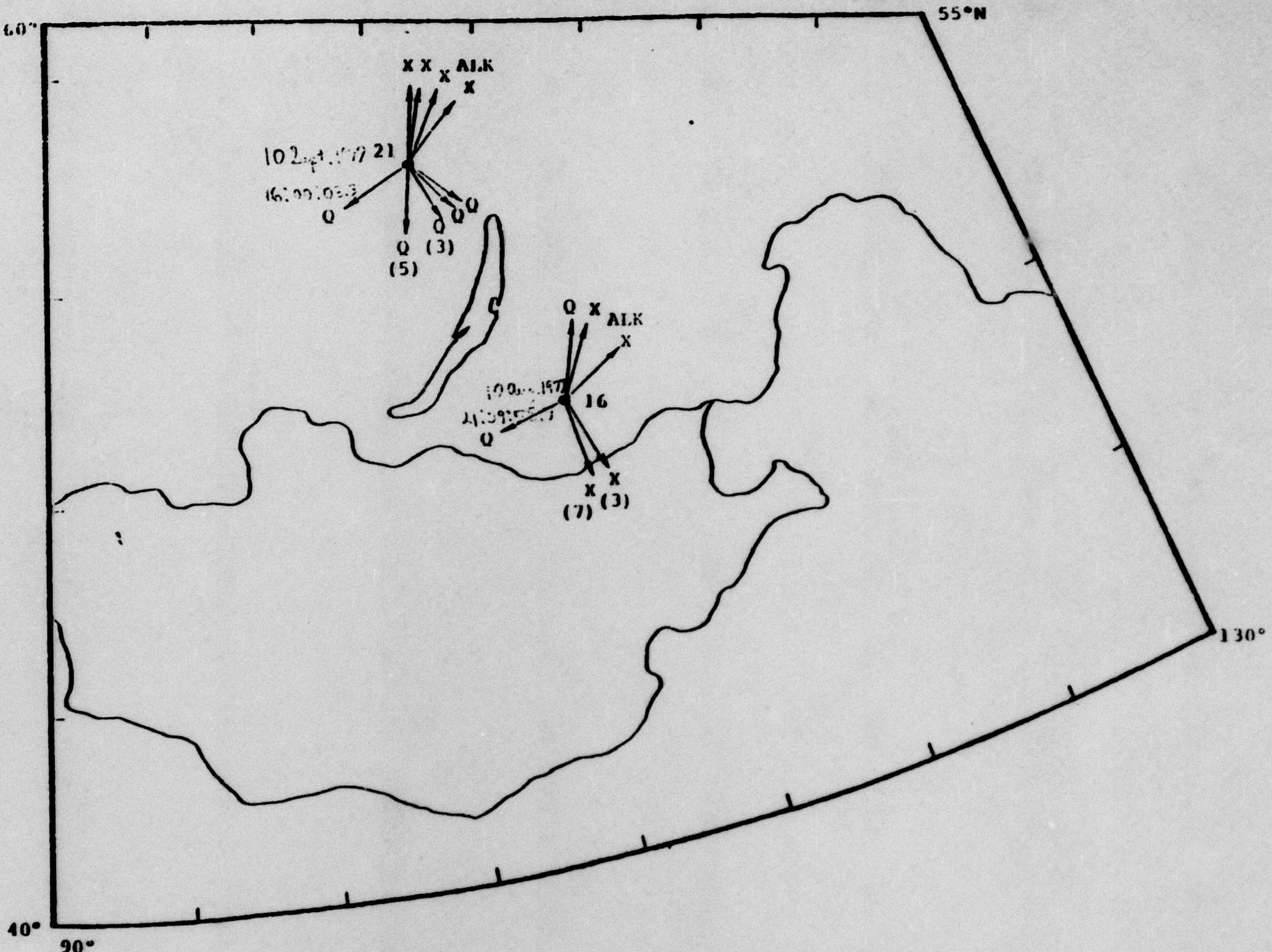
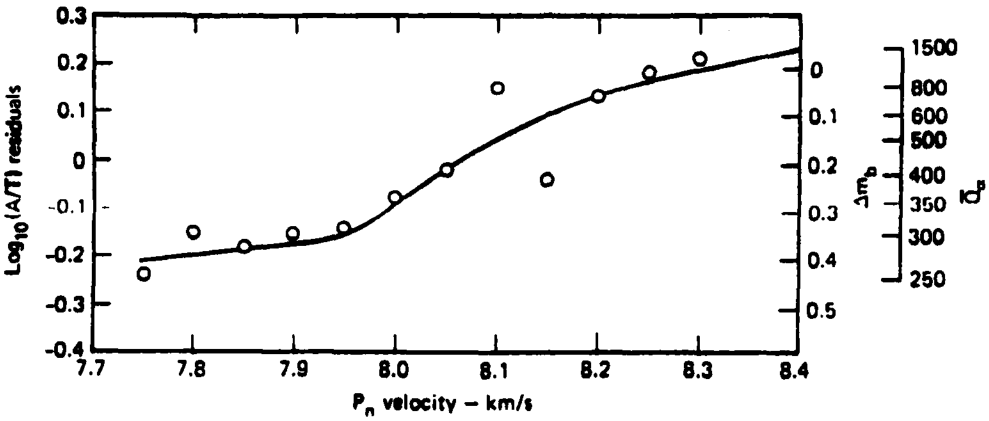


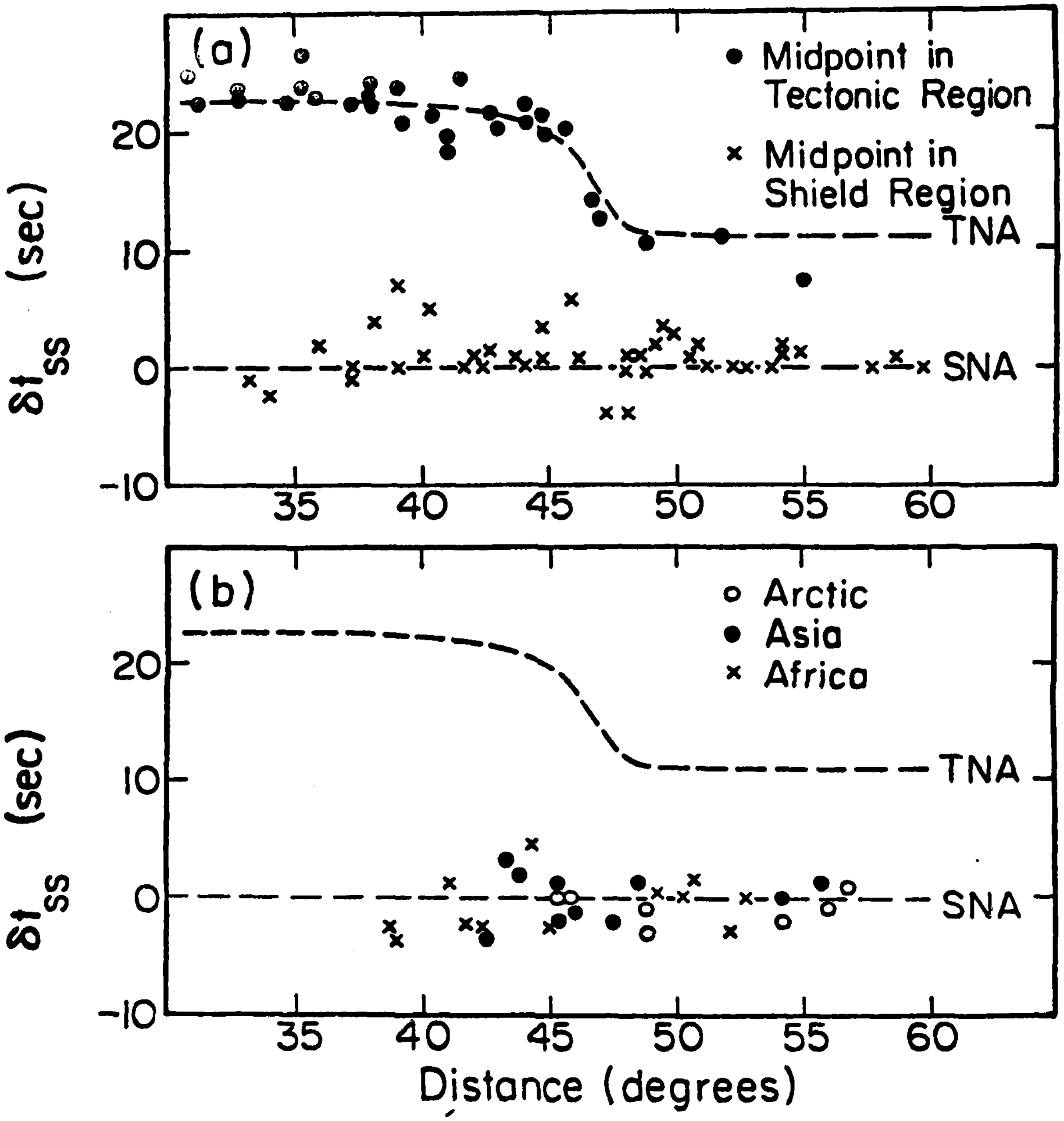
Figure 23. Map of the Lake Baikal region showing azimuthal dependence of event identification for two presumed explosions (16 and 21). Note that several Priority 1 stations are included here to augment the argument.

Slide 19.



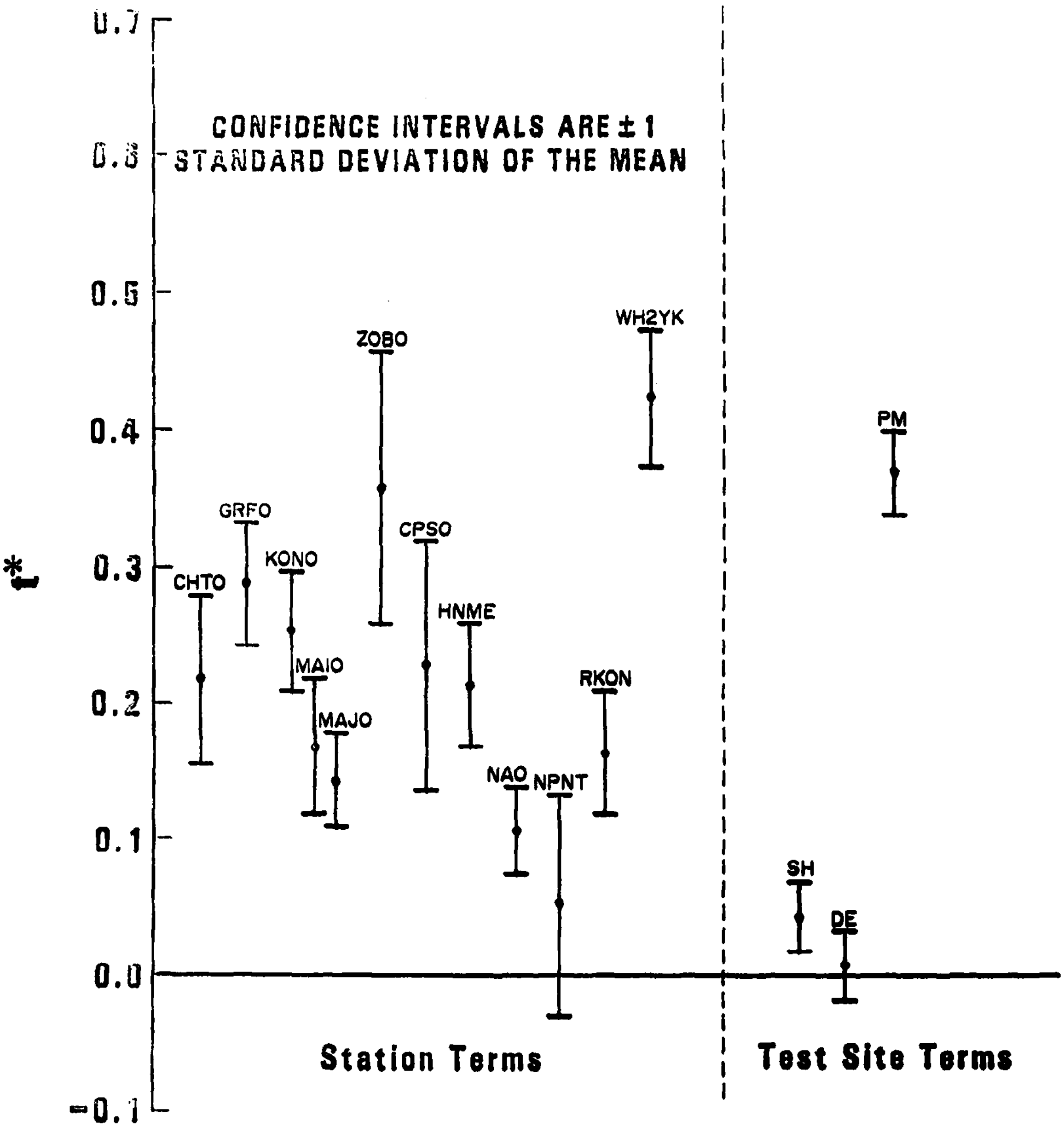
Observed relationship between P_n and $\log_{10}(A/T)$ residuals, comparison with calculated Δm_b corrections and Q_α for T = 0.75s (from Marshall et al, 1979).

Slide 20.



Plot of δt_{SS} (SS-S time residuals in seconds) versus distance for (a) North American data and (b) foreign test site data.

Slide 21.


CONFIDENCE INTERVALS ARE ± 1
STANDARD DEVIATION OF THE MEAN
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
-0.1
CHTO
GRFO
KONO
MAIO
MAJO
ZOBO
CPSO
HNME
NAO
NPNT
RKON
WH2YK
PM
SH
DE
Station Terms
Test Site Terms

Slide 18.